ground water

Movement of Water Infiltrated from a Recharge Basin to Wells

by David R. O'Leary[1,2], John A. Izbicki[1], Jean E. Moran[3], Tanya Meeth[4], Brandon Nakagawa[5], Loren Metzger[6], Chris Bonds[4], and Michael J. Singleton[7]

Abstract

Local surface water and stormflow were infiltrated intermittently from a 40-ha basin between September 2003 and September 2007 to determine the feasibility of recharging alluvial aquifers pumped for public supply, near Stockton, California. Infiltration of water produced a pressure response that propagated through unconsolidated alluvial-fan deposits to 125 m below land surface (bls) in 5 d and through deeper, more consolidated alluvial deposits to 194 m bls in 25 d, resulting in increased water levels in nearby monitoring wells. The top of the saturated zone near the basin fluctuates seasonally from depths of about 15 to 20 m. Since the start of recharge, water infiltrated from the basin has reached depths as great as 165 m bls. On the basis of sulfur hexafluoride tracer test data, basin water moved downward through the saturated alluvial deposits until reaching more permeable zones about 110 m bls. Once reaching these permeable zones, water moved rapidly to nearby pumping wells at rates as high as 13 m/d. Flow to wells through highly permeable material was confirmed on the basis of flowmeter logging, and simulated numerically using a two-dimensional radial groundwater flow model. Arsenic concentrations increased slightly as a result of recharge from 2 to 6 μg/L immediately below the basin. Although few water-quality issues were identified during sample collection, high groundwater velocities and short travel times to nearby wells may have implications for groundwater management at this and at other sites in heterogeneous alluvial aquifers.

Introduction

Recharge of groundwater by infiltration from ponds is a commonly used inexpensive approach to replenish groundwater. The approach was first used in the early 1900s to recharge coastal aquifers in California (Weeks 2002), and later in the 1930s to recharge aquifers underlying Long Island (Leggette and Brashears 1938). Since that time, recharge by infiltration from ponds has seen widespread application to unconfined aquifers because of its simplicity and low cost (Fetter 1988).

Engineering considerations related to infiltration from ponds include pond design (Rao and Sarma 1981; Pettyjohn 1981; Zomorodi 1988; Bouwer 2002), the effects of clogging (Moravcováa et al. 1968; Behnke 1969), and air entrainment (Christensen 1944; Bianchi and Haskell 1966; Heilweil et al. 2004) on infiltration rates, and mounding of water at the saturated zone (Hantush 1967; Marino 1974; Singh et al. 1976; Bouwer and Rice 1989). In early applications, comparatively shallow depths to water were favored when locating recharge ponds intended to

[1]United States Geological Survey, California Water Science Center, 4165 Spruance Road Suite 200, San Diego, CA 92101.
[2]Corresponding author: United States Geological Survey, California Water Science Center, 4165 Spruance Road Suite 200, San Diego, CA 92101; (619) 225-6157; fax: (619) 225-6101; doleary@usgs.gov
[3]California State University, East Bay, 25800 Carlos Bee Boulevard, Hayward, CA 94542.
[4]California Department of Water Resources, P.O. Box 942836 Sacramento, CA 94236.
[5]San Joaquin County Public Works, 1810 Hazelton Avenue, Stockton, CA 95205.
[6]United States Geological Survey, California Water Science Center, 6000 J Street, Sacramento, CA 95819.
[7]Lawrence Livermore National Laboratory, 7000 East Avenue, Livermore, CA 94550.
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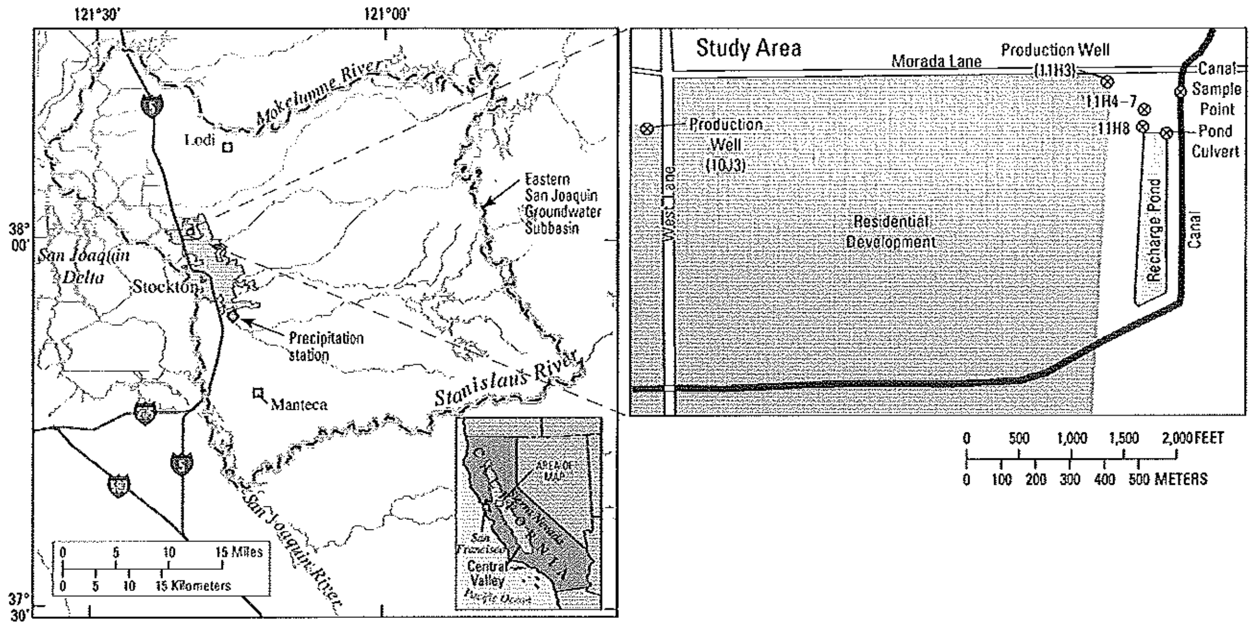

Figure 1. Location of study area, Stockton, California.

recharge water table aquifers. Although shallow depths to water reduced the time of travel to the water table, the need for a sufficiently thick unsaturated zone to allow mounding of recharge water at the water table was recognized. In recent years, the role of unsaturated zones greater than 100 m thick in the transmission and storage of infiltrated water between land surface to the water table have been increasingly studied (Ellett 2002; Flint and Ellett 2004; Izbicki et al. 2008a).

Numerous researchers have reported on the beneficial effects of infiltration through the unsaturated zone on water quality. These benefits, especially the removal of organic carbon and attenuation of microbiological contamination, have encouraged the recharge of water from a wide range of sources having potentially impaired water quality including stormflow and treated municipal waste water (Crook et al. 1990; Bouwer 1991). However, a number of studies have shown that some constituents, such as iron, manganese, and other trace elements can be mobilized from the pond bottom or shallow aquifer deposits if anoxic or suboxic conditions develop (Greskowiak et al. 2005; McNab et al. 2009).

Regulations intended to protect public health require minimum travel times be maintained from recharge ponds to extraction wells (California Department of Health Services 2008). Such regulations have prompted numerous tracer studies designed to track the movement of water infiltrated from ponds to determine the travel time from the infiltration ponds to extraction wells (Herndon et al. 2003; Clark et al. 2004). These studies have addressed the areal extent of applied water and lateral movement of water infiltrated from ponds. Only a few studies have looked closely at the physical movement of infiltrated water with depth and the movement of this water to nearby production wells (Moran and Halliwell 2003; Quast et al. 2006).

Artificial recharge is an important management tool for water suppliers, especially in aquifers where withdrawals exceed recharge, leading to water-supply shortages and water-quality degradation. The City of Stockton, California, about 130 km east of San Francisco (Figure 1), relies on groundwater for about 20% of its public supply (City of Stockton 2010). Groundwater recharge within the Eastern San Joaquin Groundwater Subbasin is about 1.11×10^9 m^3/year, and pumpage exceeds recharge by 1.85×10^8 m^3/year (GBA 2004). In the 1950s, water levels in parts of the subbasin declined to below sea level and chloride concentrations in a number of wells increased (DWR 1967). As a result, the City of Stockton is evaluating the use of local surface water and urban stormflow runoff to recharge underlying alluvial aquifers. Water infiltrated from recharge (spreading) ponds is typically less expensive and subject to fewer regulations than injection wells. The pond (Detention Basin No. 2) evaluated as part of this study is a 40-ha unlined basin constructed by the City of Stockton. Detention Basin No. 2 (herein referred to as the basin) allows the infiltration of water to the water table 15 to 20 m beneath the basin. Water infiltrated from the basin can then be withdrawn at a later date through nearby production wells.

Purpose and Scope

The purpose of this study was to evaluate the effects of artificial recharge from a pond on the quality and quantity of water in an underlying aquifer. This study is part of a larger study of the sources of high-chloride water to wells and groundwater recharge in the Eastern San Joaquin Groundwater Subbasin (Izbicki et al. 2006).

The scope of the study included drilling, monitoring well installation, continuous water level data collection, and collection of water-quality data from surface water, monitoring wells, and public-supply wells during infiltration from the basin. The study also included collection of coupled wellbore flow and depth-dependent water-quality data from a nearby production well (well 2N/6E-11H3) and development of a two-dimensional radial groundwater flow model to simulate movement of water to the production well. Results of data collection and modeling were interpreted with respect to results of a sulfur hexafluoride (SF_6) tracer test done at the site by Lawrence Livermore National Laboratory (LLNL).

Hydrogeologic Setting

The study area is within the Eastern San Joaquin Groundwater Subbasin (DWR 2006) within the San Joaquin Valley of California (Figure 1). The subbasin is underlain by several hundred meters of consolidated, partly consolidated, and unconsolidated sedimentary deposits (DWR 1967). At the study site, low permeability volcanic deposits known as the Mehrten formation (Curtis 1954), situated at approximately 200 m bls, separate overlying alluvial-fan deposits from underlying marine deposits and form the effective base of fresh water. Interspersed throughout the Mehrten formation are volcanic debris flows (lahars) (Curtis 1954). These debris flows likely have low permeability and may act as confining layers where they are aerially extensive. The volcanic deposits are blanketed by a layer of alluvium eroded from these same deposits about 30 m thick. The remainder of the overlying deposits consists of alluvial-fan deposits eroded primarily from the Sierra Nevada.

Prior to the onset of groundwater pumping, groundwater movement in the alluvial-fan deposits was from recharge areas along the foothills of the Sierra Nevada to groundwater discharge areas near the San Joaquin Delta (Mendenhall 1908). Recharge also occurred as infiltration of surface water along the upstream reaches of rivers and streams on the alluvial-fan deposits, while groundwater discharge occurred along the lower reaches of these streams (Piper et al. 1939). During the study, the regional groundwater gradient at the study site was to the southeast toward the regional pumping depression.

Under present-day conditions, groundwater recharge within the subbasin is about 1.11×10^9 m^3/year, pumping exceeds recharge by 1.85×10^8 m^3 and water levels in parts of the subbasin are declining at rates as high as 0.61 m/year (GBA 2004). Continuous water level data collected at this study site and seven other multiple-well monitoring sites in the Eastern San Joaquin Groundwater Subbasin near Stockton generally show downward gradients from the water table to major producing zones within the alluvial aquifer. The depth to the top of the saturated zone at the basin fluctuates seasonally from depths of about 15 to 20 m (Clark et al. in review).

Recharge Operations

Several detention basins, constructed for flood control, are located within the Stockton area. The city is considering using these basins for groundwater recharge during the dry summer months when they are not needed for flood control. Detention Basin No. 2, in the northern

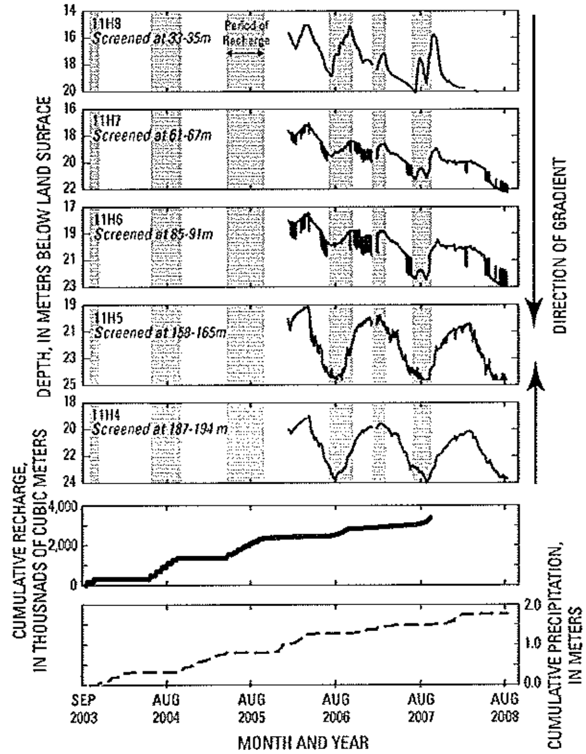

Figure 2. Water level data from wells 2N/6E-11H4-8, cumulative recharge, and cumulative precipitation near Detention Basin No. 2, Stockton, California.

portion of Stockton, was selected for study because of its proximity to a major canal, which provides a source of recharge water from the Mokelumne River (CET 2003).

Recharge at the basin began in September 2003, and continued through the fall of 2007 (Figure 2). During a typical recharge cycle, the basin was filled to a capacity of 210,000 m^3. Water was then allowed to infiltrate into the underlying alluvium. After water levels in the basin declined, the basin would then be refilled. The number of recharge cycles (filling of the basin) ranged from four to nine per season (CET 2006). Between September 2003 and September 2007, about 3.3×10^6 m^3 of water from the Mokelumne River were infiltrated from the basin to the water table aquifer (CET 2003, 2006). In addition, stormflow runoff from urban and commercial development infiltrated from the basin during the rainy season. The quantity of stormflow infiltrated from the basin was not directly measured.

Approach

The study was done collaboratively by the U.S. Geological Survey (USGS), California Department of Water Resources, and LLNL with support from the Northeastern San Joaquin Ground Water Banking Authority, the City of Stockton, the California State Water Resources Control Board, and local consultants.

Water from the Mokelumne River was first infiltrated from the basin in September 2003. Inflow to the basin and water levels in the basin were monitored by CET (2003, 2006). The USGS drilled a multiple-well site adjacent to the basin and collected water-quality data from the site in May 2005 (Clark et al. in review). Continuous water level data were initially collected from the multiple-well site by the USGS between January 2006 and July 2008 (Clark et al. in review). DWR monitored the water quality of basin inflow, the basin, and the multiple-well site quarterly for a period of 1 year to determine the effect of seasonal changes in the source of recharge water on basin and groundwater quality (DWR 2008). DWR also collected a sample of basin inflow during a storm event. LLNL applied a SF_6 tracer to the basin in October 2006 and DWR and LLNL monitored nearby wells to monitor the movement of the tracer in the subsurface. LLNL also analyzed samples for chemistry and isotopic data (Moran et al. 2009). During the tracer experiment, the USGS collected coupled wellbore flow and depth-dependent water-quality data from a nearby production well (11H3) to determine the movement of water from the basin to the well (Clark et al. in review). Wellbore flow data were simulated using a two-dimensional radial groundwater flow model and simulation results were compared with results of the tracer experiment.

Water-Quality Data

Water-quality data were collected from a multiple-well site and then from the basin and the multiple-well site on a quarterly basis from May 2006 to March 2007 to determine seasonal effects on water quality resulting from recharge and infiltration of stormflow during the rainy season.

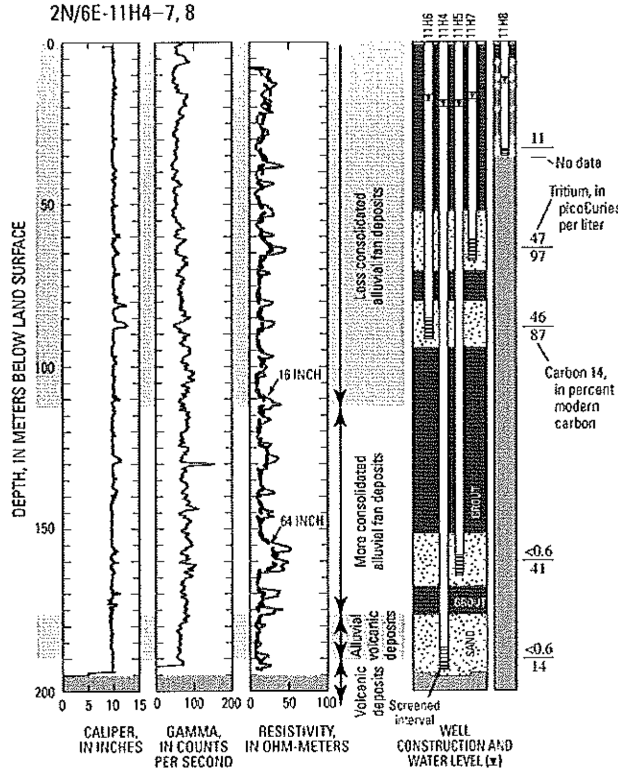

Figure 3. Selected geophysical logs and well-construction data for multiple-well site 2N/6E through 11H4-8, Stockton, California, May 2005.

Analytical Methods

Samples from the recharge basin and monitoring wells collected by the DWR were analyzed at DWRs Bryte Laboratory in Sacramento, California. Samples were analyzed for major and minor ions and trace elements, dissolved organic carbon, volatile organic compounds, glyphosphate, chlorinated organic pesticides, phosphorus and nitrogen pesticides, and chlorinated phenoxy acid herbicides (DWR 2008). These quarterly samples also were analyzed for the stable isotopes of oxygen and hydrogen by the USGS. Analyses of water samples for dissolved SF_6 were performed by LLNL (Moran et al. 2009). Sample analyses also included field parameters (pH, specific conductance, dissolved oxygen, and alkalinity), nutrients, the stable isotopes of oxygen and hydrogen (oxygen-18 and deuterium, respectively), the radioactive isotope of hydrogen (tritium), and the stable and radioactive isotopes of carbon (carbon-13 and carbon-14, respectively) (Clark et al. in review).

Results

Test Drilling and Water Level Data Collection

A multiple-well monitoring site was drilled in May 2005 using mud rotary drilling to obtain samples of geologic materials, to collect lithologic and geophysical data, and to install monitoring wells for collection of water level data and water-quality data. The monitoring well is situated approximately 70 m north of the basin and approximately 165 m southeast of production well 11H3 (Figure 1). The monitoring well was drilled to a depth of 195 m, completely penetrating the alluvial-fan deposits and the alluvial deposits eroded from the underlying volcanic deposits. The borehole partly penetrated the underlying volcanic deposits that compose the effective base of the freshwater aquifer. Drilling was stopped at this depth because of the increasing consolidation of the deposits (Figure 3). Geologic, lithologic, and geophysical data collected during drilling were described by Clark et al. (in review).

Four, 10-cm diameter, polyvinyl chloride (PVC) wells (2N/6E-11H4, -11H5, -11H6, and -11H7, from deepest to shallowest) were installed at depths of 194, 165, 91, and 67 m, respectively (Figure 3), with 6.1 m screens at the base of each well. Well 11H4 was completed near the top of the volcanic deposits in alluvium eroded from those deposits, while wells 11H5 through 7 were completed in the overlying alluvial-fan deposits. An additional borehole situated approximately 26 m north of the detention basin was drilled to a depth of 36 m and a 10-cm monitoring well (2N/6E-11H8) was installed in the alluvial-fan deposits along with several suction-cup lysimeters intended to monitor water quality at and above the water table.

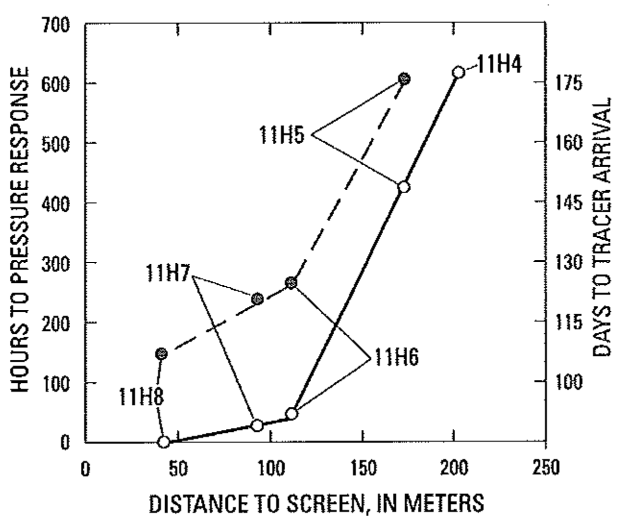

Figure 4. Water level response to basin infiltration and time to tracer arrival in wells 2N/6E-11H4-8 near Detention Basin No. 2, Stockton, California, September 2003 to August 2008.

Water levels were monitored continuously from wells 11H4 through 8 between January 2006 and December 2008 (Figure 2). Water level data was transmitted via a Geostationary Operational Environmental Satellite (GOES), providing near-real-time data from the site for management of recharge activities. Water levels in monitoring wells 11H5 through 8 decrease with depth, consistent with the regional downward hydraulic gradient in the subbasin (Clark et al. in review) (Figure 2). In contrast, water levels in the deepest well 11H4 are higher than those in well 11H5 indicating the potential for the upward movement of water at depth.

Water levels in all wells responded to seasonal pumpage. Water levels in wells 11H6 and 11H7 respond to daily pumping in nearby public-supply well 11H3 about 200 m northwest of the basin. There are seven screened intervals in production well 11H3 situated at about 79 to 82 m, 96 to 97 m, 107 to 111 m, 114 to 117 m, 128 to 130 m, 140 to 145 m, and 148 to 151 m, bls.

When water was infiltrated from the basin, pressure-head changes were observed in wells at depths up to 194 m bls. Pressure-head responses to basin infiltration were almost immediate in the shallowest well 11H8, and apparent within about 2 d after the onset of infiltration in wells 11H6 and 11H7 (Figure 4). The magnitude of the pressure-head response decreased with depth and occurred between 16 and 25 d after the onset of infiltrations in the two deeper wells, 11H4 and 11H5 (Figure 4)—possibly reflecting increasing consolidation and clay content and decreasing permeability of deposits deeper than about 125 m bls. The magnitude of the pressure-head response for well 11H8 was about 1 to 2 m; wells 11H7 and 11H6 were about 0.5 to 1 m; and wells 11H5 and 11H4 was less than 0.5 m. Although subtle changes are present in geophysical logs near this depth (Figure 3), the increased consolidation of the alluvial-fan deposits with depth was not obvious on the basis of geologic or test drilling data.

Chemistry of Recharge Water and Water from Wells

During quarterly sampling from May 2006 to March 2007, water in the basin was generally low in dissolved solids with concentrations ranging from 56 to 129 mg/L. The quarterly sample collection period included the winter rainy season when recharge from infiltration of stormflow occurred (Figure 2). Dissolved-solids concentrations were higher during the winter rainy season when stormflow runoff was present in the basin, and concentrations were lower when Mokelumne River water was present in the basin (DWR 2008). Dissolved organic carbon (DOC) concentrations in water from the basin typically ranged from 4.2 to 7.8 mg/L. However, one sample collected from stormflow runoff influent to the basin had a DOC concentration of 18 mg/L. Low concentrations of chlorinated phenoxy herbicides (2,4-D, dicambia, and M-chlorophenylpiperazine [MCPP]) not exceeding 0.6 μg/L were present in basin water. Similarly, low concentrations of diazinon and trifurlan not exceeding 0.1 μg/L also were present in stormflow influent to the recharge basin (DWR 2008). Winter rainy season and stormflow dissolved organic carbon concentrations and pesticide concentrations are consistent with the quality of stormflow used for groundwater recharge in urban areas elsewhere in California (Schroeder 1995; Izbicki et al. 2004b, 2007).

Dissolved-solids concentrations from wells in the multiple-well monitoring site adjacent to the basin were generally low, although slightly higher than concentrations in the basin, with concentrations ranging from 126 to 325 mg/L (DWR 2008). Dissolved-solids concentrations were lowest in the shallowest well closest to the basin. Changes in dissolved-solids concentrations with depth were accompanied with a shift in major-ion composition from calcium bicarbonate-dominated to sodium bicarbonate-dominated water in the deepest well 11H4 (DWR 2008). Similar changes in major-ion chemistry with depth have been observed in alluvial aquifers elsewhere in California and are consistent with a combination of cation exchange, and calcite precipitation as a result of increases in alkalinity from oxidation of organic material, and in some cases sulfate reduction (Izbicki et al. 1998). Similar changes in the chemistry of water infiltrated from basins to recharge underlying aquifers have been observed elsewhere in the subbasin (McNab et al. 2009).

Dissolved organic carbon concentrations in water from monitoring wells ranged from 0.5 to 1.5 mg/L and were lower than concentrations in the basin and stormflow influent to the basin. Dissolved organic carbon concentrations decreased with depth with the higher concentrations ranging from 1.1 to 1.5 mg/L in the shallowest well 11H8, 36 m bls.

Reducing conditions with dissolved concentrations of 0.5 mg/L or less were observed in the shallowest monitoring well (11H8). Oxic conditions, characterized by dissolved oxygen concentrations as high as 3.6 mg/L, prevailed in wells 11H6 and 11H7, at 91 and 67 m bls, respectively. Reducing conditions and dissolved oxygen concentrations less than 0.5 mg/L also were present in the deeper wells at the site 11H4 and 1H5, at 194 and 165 m bls, respectively.

Arsenic was present at concentrations between 1 and 2 µg/L in water from the basin and stormflow influent to the basin. Arsenic concentrations in the shallowest well at this site, 11H8, ranged between 4 and 6 µg/L and were greater than concentrations in either the recharge basin, stormflow influent to the basin, or concentrations in wells 11H6 and 11H7. Arsenic was present at concentrations as high as 30 µg/L in the two deepest wells at the site, 11H4 and 11H5 (DWR 2008; Clark et al. in review). Arsenic concentrations in water from these wells exceeded the U.S. Environmental Protection Agency Maximum Contaminant Level of 10 µg/L. Izbicki et al. (2008b) showed that arsenic concentrations as high as 60 µg/L in water from wells deeper than about 100 m in the Eastern San Joaquin Groundwater Subbasin were the result of subsurface geology and naturally occurring reducing conditions at depth.

Chlorinated phenoxy acid herbicides, diazinon, and triflualin, present at low concentrations in the recharge basin and in stormflow runoff influent to the basin, were not detected in water from the multiple-well site—consistent with sorption or degradation of these compounds. However, simazine was present at a concentration of 0.02 µg/L in one sample from the shallowest well at the site (DWR 2008).

Isotopic Composition of Recharge Water and Water from Wells

Oxygen-18 and deuterium (δ^{18}O and δD, respectively) are naturally occurring stable isotopes of oxygen and hydrogen, respectively. Most of the world's precipitation originates from the evaporation of seawater. As a result, the δ^{18}O and δD composition of precipitation and most groundwater is linearly correlated and distributed along a line known as the global meteoric water line (Craig 1961). The δ^{18}O and δD composition of a water sample relative to the meteoric water line and relative to the composition of water from other areas can provide a record of the source and evaporative history of the water. δ^{18}O and δD data were used to evaluate the movement of water infiltrated from the basin. As part of this analysis, samples from the basin and from nearby monitoring wells were compared with data from other wells in the Eastern San Joaquin Groundwater Subbasin collected by Izbicki et al. (2006).

The δ^{18}O and δD composition of water collected from the recharge basin ranged from −6.3 to −11.2‰ and −42 to −81‰, respectively. Stormflow water, derived from local precipitation, had a heavier (less negative) isotopic composition than water from the Mokelumne River, derived as runoff from the higher altitudes of the Sierra Nevada (Figure 5). Water from the two deepest monitoring wells at the site, 11H4 and 11H5, had δ^{18}O and δD compositions ranging from −8.4 to −9.1‰ and −60 to −65‰, respectively (Figure 5). These samples were near the average isotopic composition of stormflow and water from wells in the Eastern San Joaquin Groundwater

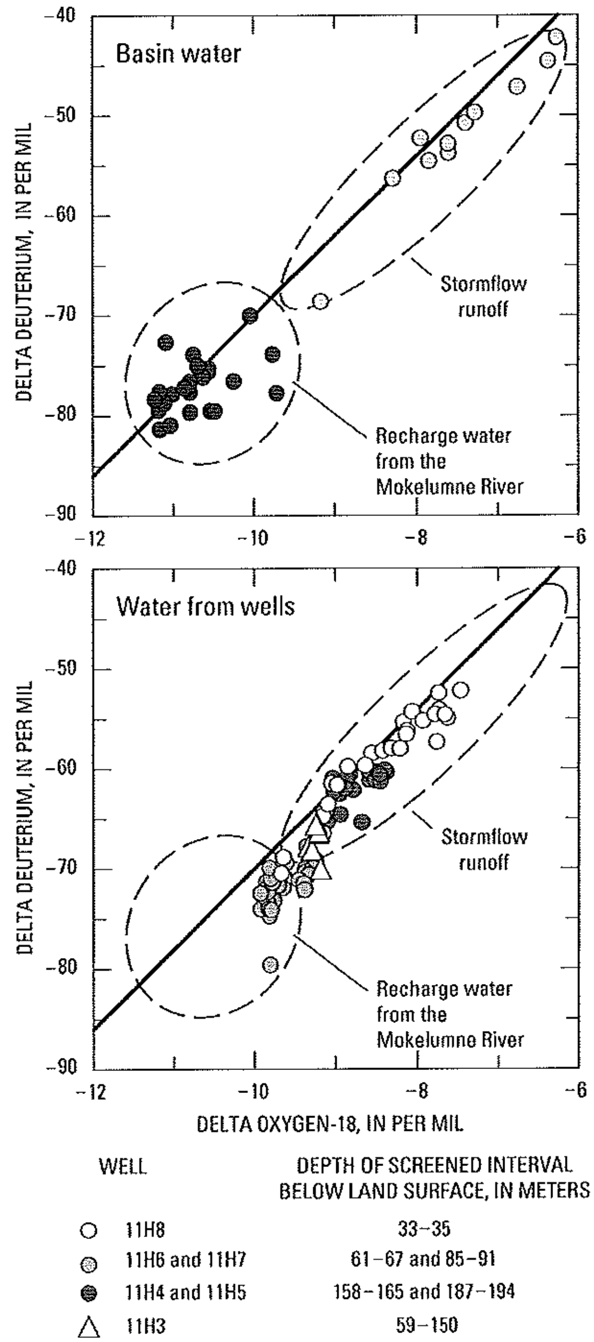

Figure 5. Delta Oxygen-18 and delta deuterium composition of water from Detention Basin No. 2, the adjacent multiple-well site 2N/6E-11H4-8, and nearby public-supply well 2N/6E-11H3, Stockton, California, May 2005 to June 2007.

Subbasin (Izbicki et al. 2006). Water from monitoring wells, 11H6 and 11H7, had δ^{18}O and δD compositions ranging from −9.2 to −9.9‰ and −78 to −80‰, respectively (Figure 5). These values were similar to the isotopic composition of recharge water from the Mokelumne River suggesting that some of the water came from the basin and had infiltrated to this depth (Izbicki et al. 2006). In contrast, water from the shallowest monitoring well 11H8 had a wide range in δ^{18}O and δD compositions ranging from −6.3 to −9.6‰ and −42 to −70‰, respectively (Figure 5). Not unexpectedly, the $\delta^{18}O$ and δD composition of water from this well varied depending on the source of water in the basin. The isotopically light water reflects the relatively light composition of Mokelumne River water, which was the largest source of water to the basin. The isotopically heavy water reflects the heavier composition of locally derived stormflow.

Tritium, a radioactive isotope of hydrogen having a half-life of about 12.3 years, was used to indicate the presence of recent water. Water containing tritium was isolated from the atmosphere after the advent of atmospheric testing of nuclear weapons in 1952, water not containing measurable tritium was isolated from the atmosphere prior to 1952. Tritium data show recent water containing tritium activities between 11 and 47 pCi/L in water from wells 11H6, 11H7, and 11H8 consistent with a recent recharge of the water (Figure 3). The deeper wells at the site had tritium activities less than 0.6 pCi/L indicating that those wells do not receive recent recharge.

Carbon-14, a radioactive isotope of carbon having a half-life of about 5730 years, was used to indicate the presence of older groundwater. Carbon-14 activities can approach, or exceed, 100 pmC for recently recharged water containing tritium. Carbon-14 activities are lower for older groundwater isolated from the atmosphere for long periods of time. For example, neglecting reactions between water and aquifer minerals, water having a carbon-14 activity of 50 pmC has been isolated from the atmosphere for about one half-life, or about 5730 years. Carbon-14 activities in water from wells 11H4 and 11H5 were 14 and 41 pmC, respectively (Figure 3). These carbon-14 activities are consistent with older groundwater (uncorrected carbon-14 ages of about 16,000 and 7000 years before present, respectively) and suggest little interaction between the shallow and deeper groundwater. In contrast, carbon-14 activities from the shallower wells 11H6 and 11H7 were 87 and 97 pmC, respectively (uncorrected carbon-14 ages of about 1000 and 170 years before present, respectively). As stated above, samples from these wells also contain more negative $\delta^{18}O$ and δD compositions consistent with the presence of recently recharged water from the Mokelumne River (Figure 3).

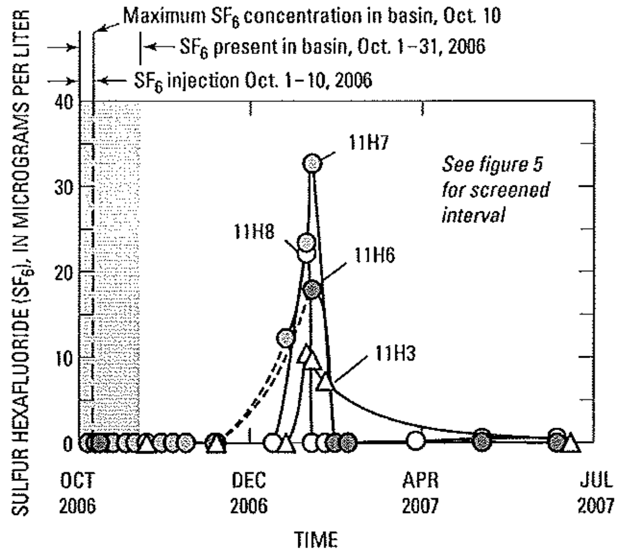

Figure 6. Sulfur hexafluoride (SF_6) tracer concentrations in water from Detention Basin No. 2, the adjacent multiple-well site 2N/6E-11H4-8, and nearby public-supply well 2N/6E-11H3, Stockton, California.

Tracer Test Data

A tracer test, using SF_6, was done to determine the rate of movement of water from the basin to nearby production wells (Moran et al. 2009). SF_6 is a poorly soluble, easily measured, nonreactive gas. Although very low concentrations of SF_6 (10^{-15} mol/L; 1.46×10^{-6} µg/L) may occur naturally in some aquifers (Deeds 2008), SF_6 is an excellent tracer of the movement of water at concentrations commonly used in tracer tests. SF_6 is nontoxic and permission to add small concentrations to the recharge water was obtained from the California State Department of Public Health Services.

The tracer was introduced into the basin between October 3 and 10, 2006 by diffusing SF_6 gas into the water at a flow rate of 20 cm^3/min. About 61,675 m^3 (50 ac-ft) of water was delivered to the basin during this period, with a SF_6 injection rate of 20 cm^3/min, for an average SF_6 concentration of about 40 µg/L. In practice, the total mass of tracer dissolved into the water is difficult to estimate because dissolution of the gas is a function of the water temperature and gas diffusion at the basin surface. The nearby monitoring and production wells were sampled over the following 8 months (Moran et al. 2009). Velocities based on tracer first arrival times represent the fastest flow and may not reflect the average groundwater flow velocity, which may be slower because of dispersion and lithologic heterogeneity (Freeze and Cherry 1979).

Samples were collected periodically in multiple-well site 2N/6E-11H4-8 and nearby public-supply well 2N/6E-11H3 to estimate the combined vertical and horizontal travel time for water to move from the basin to the sampled wells (Figures 4 and 6). Detectable SF_6 was measured in well 11H7 (61-67 m bls) 108 d after the onset of injection, in well 11H8 (33-35 m bls) 119 d after the onset of injection, in well 11H7 (61-67 m bls) 108 d after the onset of injection, in well 11H6 (85-91 m bls) 122 d after the onset of injection, and in well 11H5 (158-165 m bls) 175 d after the onset of injection. SF_6 was not detected in the deepest well at the site, 11H4 (187-194 m bls), during the experiment, consistent with water level data that show an upward hydraulic gradient at this depth.

SF_6 was first detected in public-supply well 11H3 119 d after the onset of injection yielding an average horizontal tracer velocity of about 1.7 m/d. The linear vertical tracer velocities between wells 11H5 and 11H6 and the basin were about 0.3 and 0.5 m/d, respectively. The linear horizontal tracer velocity between monitoring well 11H7 and public-supply well 11H3 was around 13 m/d. The tracer arrived at production well 2N/6E-10J3, 1560 m west of the basin, 169 d after the onset of injection, and the linear horizontal tracer velocity between the basin and well 10J3 was about 9.2 m/d.

Coupled Wellbore Flow and Depth-Dependent Water-Quality Data

Coupled wellbore flow and depth-dependent water-quality data were collected from public-supply well 11H3 in June 2007, to determine the distribution of flow and water quality in the long-screened well. It was not possible to collect these data earlier in the tracer test because of a pump malfunction in the production well. Wellbore flow data were collected using a commercially available impellor (spinner) tool and depth-dependent water-quality data were collected using a 2.5-cm-diameter, gas-displacement pump (Izbicki 1999, 2004a). At the time of collection, the static water level in 11H3 was about 17 m bls. The well was pumped at its production rate of 54 L/s for about 10 h with a measured drawdown of 51 m.

The wellbore flow log data show that flow into well 11H3 is not uniformly distributed through the well screens (Figure 7). About 70% of the flow into well 11H3 entered through two screened sections at 107-111 and 114-117 m bls. Only 11% of the yield to the well was contributed from the three screens completed in the deeper deposits below about 125 m (Figure 7). These data, illustrating the distribution of flow into the well, reflect the combined effects of the heterogeneous nature of the alluvial deposits and well efficiency.

Depth-dependent water-quality data show relatively uniform chemistry in well 11H3 with depth. However, the calculated $\delta^{18}O$ and δD composition of water entering well 11H3 is isotopically lighter in samples collected from the uppermost screens between 79-82 and 96-98 m bls and is similar to water from the Mokelumne River recharged from the basin (Figure 7).

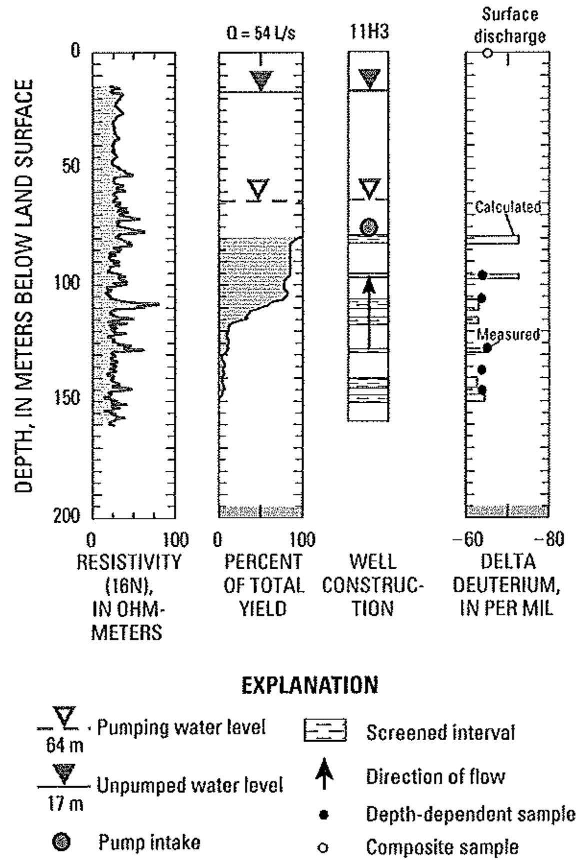

Figure 7. Wellbore flow and depth-dependent isotopic data from public-supply well 2N/6E-11H3, near Detention Basin No. 2, Stockton, California, June 20, 2007.

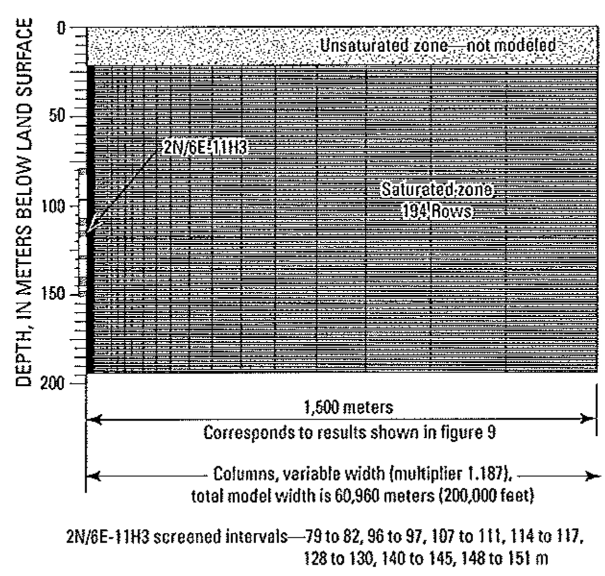

Figure 8. Model grid and boundary conditions used to simulate flow to well 2N/6E-11H3, near Detention Basin No. 2, Stockton, California.

Numerical Modeling

Groundwater flow and particle paths were simulated around well 11H3 using the computer program AnalyzeHOLE (Halford 2009) to help evaluate the effect of aquifer heterogeneity on groundwater movement and travel times. AnalyzeHOLE simulates wellbore flow using an axisymmetric, radial geometry in a two-dimensional MODFLOW model (Harbaugh et al. 2000). The model consists of one layer having 194 horizontal rows (depth) and 80 vertical columns (distance from the pumping well), and represents a cylinder of aquifer material having a radius of 60,960 m and a thickness of 194 m. The well (11H3) was simulated as a high hydraulic conductivity (K) zone (30,480 m/d) in the first column of the model (Figure 8). The well casing and well screens were simulated in the second column of the model. The well casing was simulated with a K of 0 m/d. The well screens were initially assumed to be 100% efficient and set to the same K value as the well. Well screen depths were adjusted slightly to conform to the model grid. The gravel pack and sanitary seal were simulated in the third column of the model. The gravel pack was simulated with a K of 91 m/d from the bottom of the well to the bottom of the sanitary seal which was simulated with a K of 0 m/d.

Aquifer deposits were assumed to be radially symmetric, flat-lying, and laterally extensive through the

Table 1
Details of Radial Groundwater Flow Model Construction

Data	Value	Data		Value
Spatial discretization		Assigned hydraulic conductivities of model aquifer		
Grid dimensions	194 m thick by	17–35 m	Sand/silt/clay	0.76 m/d
	60,960 m wide	35–50 m	Clay	0.0003 m/d
Number of layers	1	50–52 m	Sand/silt	1.8 m/d
Number of rows	194	52–78 m	Sand/silt/clay	0.76 m/d
Thickness of rows	1 m	78–81 m	Sand/silt	1.8 m/d
Number of columns	80	81–91 m	Silt/clay	0.08 m/d
Size of columns	Variable	91–95 m	Clay	0.0003 m/d
Column 1 (well)	0.25 m	95–99 m	Fine sand	4.0 m/d
Column 2 (casing)	0.013 m	99–107 m	Clay	0.0003 m/d
Column 3 (gravel pack)	0.13 m	107–112 m	Coarse sand	10 m/d
Columns 4–80	Multiplier 1.555	112–113 m	Clay	0.0003 m/d
		113–119 m	Fine/medium sand	6.1 m/d
Side boundary condition	No flow	119–128 m	Sand/silt/clay	0.76 m/d
Bottom boundary condition	No flow	128–130 m	Fine sand	4.0 m/d
Upper boundary condition	Water table	130–140 m	Silt/clay	0.08 m/d
(Initial water level 17 m bls)		140–145 m	Sand/silt/clay	0.76 m/d
		145–148 m	Fine sand	4.0 m/d
Hydraulic properties		148–150 m	Sand/silt/clay	0.76 m/d
Porosity	0.2	150–155 m	Fine sand	4.0 m/d
Specific storage	7.62E-07 /m	155–186 m	Sand/silt	1.8 m/d
Anisotropy	0.5	186–188 m	Clay/silt/lahar	0.003 m/d
Hydraulic conductivity (calibrated)		188–194 m	Fine sand	4.0 m/d
Well casing	0 m/d	Temporal discretization		
Clay	0.0003 m/d	Stress periods		1
Clay/silt/lahar	0.003 m/d	Length of stress period		999 d
Silt/clay	0.08 m/d	Time steps		25
Sand/silt/clay	0.76 m/d	Time step multiplier		1.2
Sand/silt	1.8 m/d	Initial time step		2.1166 d
Fine sand	4.0 m/d	Pumping rate		54 L/s
Fine/medium sand	6.1 m/d			
Coarse sand	10 m/d			
Gravel pack	91 m/d			
Well	30,480 m/d			

model domain. Hydraulic properties from literature derived values (Freeze and Cherry 1979) were initially assigned on the basis of lithologic and geophysical logs and were adjusted during model calibration.

The radial extent of the model was larger than the influence of simulated pumping from the well and no flow boundaries were used to represent the outside and the bottom of the cylinder. Regional groundwater flow and regional pumping effects were not simulated, and pumping stress from the simulated well (11H3) were assumed to dominate groundwater flow within the system. Similarly, the infiltration of water from the recharge basin was not modeled, and water extracted from storage within the aquifer was the only source of water to the well. Additional details of model construction, including temporal discretization, are summarized in Table 1.

The computer program MODPATH (Pollack 1994) is used in AnalyzeHOLE to track the movement of particles tracing groundwater flow within the model (Halford 2009). Water table surface under pumping conditions was approximated using the Theis equation. The computed water table served as the upper model boundary. Specific storage was used rather than specific yield (Table 1). The error associated with this approximation is believed to be small because differences in the simulated and actual water table configurations are small (Halford 2000; Clemo 2002). The results from this program represent groundwater-travel times and pathlines for advective transport within the model domain.

Numerical Model Calibration

The model was calibrated by adjusting aquifer hydraulic properties and comparing simulation results to measured drawdown and wellbore flow data collected from well 11H3 under pumping conditions. Model output was also compared to measured water levels in monitoring wells 11H4-8 (Figure 9).

The calculated aquifer transmissivity at the screened intervals of well 11H3, 91 m^2/d, was initially estimated

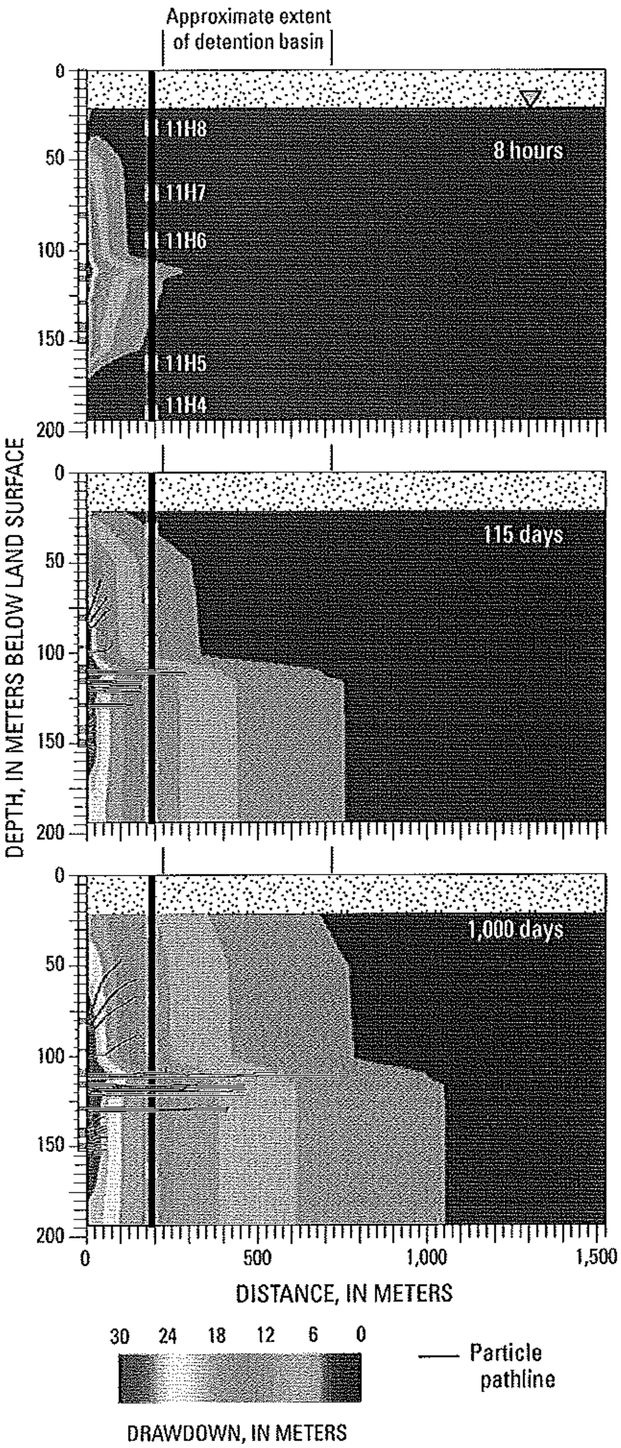

Figure 9. Simulated water level declines and inferred particle paths after 8 h, 115 d, and 1000 d of pumping from public-supply well 2N/6E-11H3, near multiple-well site 2N/63-11H4 through 8 and Detention Basin No. 2, Stockton, California.

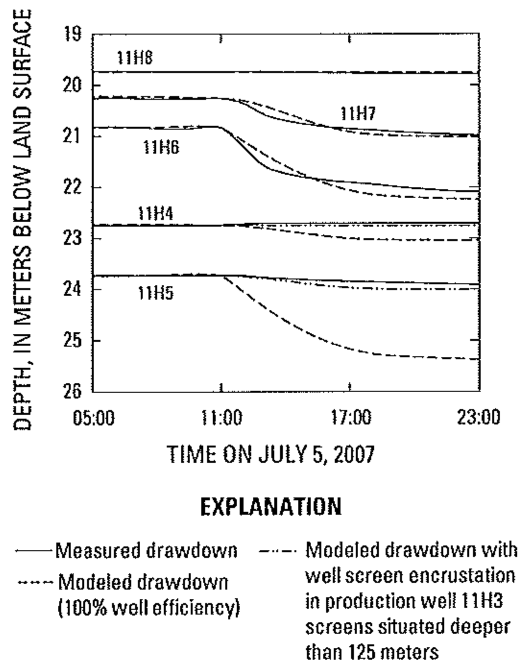

Figure 10. Measured and simulated water level declines in multiple-well site 2N/6E-11H4-8 in response to pumping (about 54 L/s) in public-supply well 2N/6E-11H3, near Detention Basin No. 2, Stockton, California.

from measured drawdown and pumpage data using a relation between specific capacity (drawdown per unit pumpage) and transmissivity (Thomasson et al. 1960). The hydraulic conductivities of lithologic units were adjusted so that the simulated transmissivity of the screened section of the model domain equaled the estimated transmissivity thereby maintaining a reasonable match between simulated and measured drawdown in well 11H3.

Initial hydraulic conductivities assigned to model layers below 125 m based on lithologic and geophysical logs allowed too much flow into the well through the deeper screens. One method of matching wellbore flow data during calibration was to decrease the hydraulic conductivity of the sand and silt units from deposits deeper than 125 m to low values similar to those typical of a clay. However, this method was rejected because the low hydraulic conductivities did not correlate with what would be expected at these depths based on the lithology observed in drill cuttings and geophysical logs. Additionally, the low hydraulic conductivities did not correspond to interpretations of regional hydrology based on water level measurements and pumping and recovery rates from monitoring wells screened at similar depths. Instead effects of screen encrustation were simulated by varying the hydraulic conductivity of the gravel pack in column 3 for screened sections below 125 m. Houben (2006) demonstrated that encrustations extend into the gravel pack opposite zones contributing flow to the well. This method proved to be effective in matching simulated wellbore flow with observed values (Figure 7). After 8 h of pumping, simulated pressure responses at locations in the model grid corresponding to the locations of well screens 11H5, 11H6, 11H7, and 11H8 were consistent with water level changes observed in the monitoring wells during a typical pumping cycle (Figure 10). Simulated pressure responses in well 11H4 were about 0.3 m greater than the measured drawdown of less than 0.1 m in response to short-term pumping cycles (less than 12 h).

The addition of a low K layer in the model at a depth roughly corresponding to the observed contact between the base of the volcanically derived alluvium and the top of volcanic deposits (Mehrten formation) decreased simulated pressure responses in well 11H4 to less than 0.1 m (Figure 10), demonstrating that deeper aquifers could be isolated from shallower aquifers by low permeability clays or the presence of lahars, which are reported to be present in the volcanic deposits (Curtis 1954).

Model calibrated hydraulic conductivities are presented in Table 1. The calibrated transmissivity of 91 m^2/d matched the initial estimated transmissivity. The calibrated model drawdown for simulated pumping of 54 L/s (850 gpm) was 49 m, slightly less than the measured drawdown of 51 m.

Numerical Model Results

Model results provide a picture of pressure responses (drawdown) and movement of water (illustrated as particles) to production well 11H3 and the surrounding aquifer (Figure 9). As noted, after 8 to 12 h of simulated pumping (approximately representative of a daily pumping cycle) simulated drawdowns in the screened intervals of shallow wells 11H4 through 11H8 were similar to observed drawdowns (Figure 10) in response to pumping in well 11H3.

After 115 d of continuous simulated pumping, approximately representative of a seasonal pumping cycle, water level declines were simulated at depths sampled by all wells at multiple-well site 11H4-8 (Figure 9). In contrast to the 8-h simulation, water levels declines were greater for the depths sampled by wells 11H4 and 11H5 in the deeper more consolidated deposits—consistent with large seasonal variations in measured water level data from these wells (Figure 2). Simulated drawdown at all depths after 115 d of simulated pumping was greater than seasonal drawdown measured at multiple-well site 11H4-8 (Figure 2) because well 11H3 is not pumped continuously during the summer pumping season.

Longer modeled pumping periods (1000 d of simulated pumping) show continued drawdown with increased water level declines in the deeper more consolidated deposits (Figure 9). The simulated distribution of water level declines with depth is consistent with the regional downward hydraulic gradient measured at multiple-well sites throughout the Eastern San Joaquin Groundwater Subbasin by Clark et al. (in review).

Particle tracking results show rapid movement of particles (water) through simulated high-conductivity units at 106-110 m bls (Figure 9). Simulated average horizontal particle velocities between public-supply well 11H3 and the basin were about 1.9 m/d. This velocity is consistent with the SF_6 tracer results that show average groundwater velocities of about 1.7 m/d between the public-supply well and the basin. Simulated horizontal particle velocities between well 11H7 and public-supply well 11H3 were about 6.7 m/d. This high velocity is consistent with the SF_6 tracer results that show groundwater velocities as high as 13 m/d between monitoring well 11H7 and public-supply well 11H3. The difference between these values probably results from the radial nature of the groundwater flow simulation and the lenticular nature of the deposits.

Particle-tracking results also show slower movement of water through deposits that overlie the high-permeability units encountered by well 11H3 (Figure 9). Simulated rates of particle movement in these shallower deposits were on the order of 0.6 m/d. This value compares well with the movement of water calculated from the SF_6 tracer data.

Numerical Model Limitations

The two-dimensional radial groundwater flow model developed to simulate wellbore flow data from well 11H3 is a simplified representation of the groundwater flow system near the well. The flatly lying aerially extensive aquifer materials simulated within the model domain were not intended to accurately represent subsurface geology, including the areal extent and hydraulic connections between these materials. Particle velocities are averaged over the distance traveled. The vertical movement of the tracer to the deeper units illustrates that the overlying low permeability units are not laterally continuous as simulated in the model. Similarly, the water table does not accurately represent regional groundwater flow or interactions between pumping wells and the recharge basin. However, this simplified model does provide a simple tool to evaluate the effects of aquifer heterogeneity on the movement of water infiltrated from the basin to well 11H3.

Discussion and Conclusions

Water from the Mokelumne River and stormflow infiltrated from the basin (Detention Basin No. 2) was of generally high quality. However, degradation of dissolved organic carbon between the basin and the shallowest well, 11H8, resulted in reducing conditions within the aquifer at this depth and dissolved oxygen concentrations of 0.5 mg/L or less, contributing to increased arsenic concentrations in well 11H8.

Application of water to the basin resulted in pressure responses throughout the alluvial deposits to a depth of 165 m bls. The pressure responses were immediate near the water table, and occurred within 2 d at depths of about 90 m. The pressure responses were smaller and occurred as late as 25 d after the onset of infiltration in deeper, more consolidated deposits.

There was a large difference in the $\delta^{18}O$ and δD composition of water from the Mokelumne River and local stormflow water infiltrated from the basin. Movement of isotopically lighter Mokelumne River water infiltrated from the basin was observed to depths as great as 91 m bls. Groundwater samples collected beneath the basin contained tritium to depths as great as 91 m, indicating recently recharged water; whereas, samples collected at greater depths do not contain measurable tritium and have uncorrected carbon-14 ages in excess of 5000 years before present. SF_6 tracer data indicate that it takes about 122 d for water to infiltrate from the basin to the highly permeable deposits at 91 m bls at the nearby multiple-well monitoring site. Once water reached these highly permeable deposits it moved rapidly to nearby pumping wells at rates as high as about 13 m/d. Both pressure responses and tracer travel times travel times between the basin and wells demonstrate the relative isolation of deposits situated deeper than well 11H6 (Figure 4), although tracer data indicate that infiltrated water reached depths as great as 165 m.

Flowmeter logging and numerical modeling in a nearby public-supply well show as much as 70% of the yield to the well was from screens installed at 107-111 and 114-117 m bls. Numerical flow modeling shows groundwater velocities within these highly permeable deposits beneath the basin are about 6.7 m/d. Although conceptually similar, this value is about half of the groundwater flow velocity obtained from the SF_6 tracer test (13 m/d). The difference probably results from simplifications associated with the two-dimensional radial flow model. It is likely that the units within the aquifer are lenticular rather than aerially extensive as assumed in the model.

Artificial recharge from surface basins is an effective method to recharge alluvial aquifers, producing water level responses and physical movement of water to depths pumped by wells. However, heterogeneity that results in high groundwater flow velocities may have implications for groundwater management; for example, highly permeable layers within the aquifer may result in unexpectedly short travel times from recharge basins to nearby pumping wells and pumping wells may induce vertical gradients increasing the downward movement of water. This may be a cause for concern in areas where long residence times are required for recharge of stormflow water, or water from other sources, to meet regulatory requirements designed to protect public health.

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